# PALEOMAGNETISM OF ANTARCTIC ACHONDRITES (II)

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***Abstract***: The paleointensity ($F_p$) of NRM of the TRM origin can be approximately estimated by comparing the NRM-lost obtained by the stepwise AF-demagnetization of NRM with the ARM-lost obtained by the same AF-demagnetization of the saturated ARM acquired in a magnetic field $h_A$ through a possible relation of (NRM-lost)/$F_p = f_0$(ARM-lost)/$h_A$, where NRM is assumed to be identical to a total TRM acquired in $F_p$, and the coefficient $f_0$ must be experimentally determined.

For ALH-77302 eucrite, $F_p$ can be determined by the method as $F_p = 0.110/f_0$ Oe. However, an experimental determination of $f_0$ for meteorites is associated with much difficulty caused by chemical and structural alterations of ferromagnetic Fe-Ni metals at elevated temperatures. A three step heating procedure which is a modification of the original TAYLOR's technique (1979) is adopted in the present study with a satisfactory success of not altering the magnetic microscopic coercivity spectrum by heating up to 790°C. The procedure consists of the first heat treatment to rid a meteorite sample of $H_2O$ vapour and loosely-bound gases at 110°C in a low vacuum space for an hour, the second one to take out tightly-bounded gases at 355°C in a high vacuum space with the help of the Ti-metal getter to absorb gases for an hour and a half, and then heating to any higher temperature to give pTRM or total TRM on the sample in the same high vacuum system.

The $f_0$ value of ALH-77302 thus obtained is $f_0 = 6.80$, and consequently $F_p$ of ALH-77302 eucrite has been obtained as $F_p = 0.016$ Oe. $F_p = 0.021$ Oe for another test specimen of the same eucrite. The paleointensity of ALH-77302 eucrite is therefore approximately 0.02 Oe.

## 1. Introduction

The paleointensity of stony meteorites have been studied with various kinds of stony meteorites (STACEY *et al.*, 1961; WEAVING, 1962; GUS'KOVA, 1963; BANERJEE and HARGRAVES, 1972; BUTLER, 1972; GUS'KOVA, 1972; NAGATA and SUGIURA, 1977). As summarized by NAGATA (1979a), it seems that the paleointensity of Allende C3 chondrite is about 1.0 Oe as summary of experimental results obtained by four different research groups with the aid of different experimental methods, while the paleointensity of several achondrites has been determined as 0.05–0.2 Oe, the average value being about 0.1 Oe (NAGATA, 1979a, 1980a).

Since the stable component of natural remanent magnetization (NRM) of chondrites and achondrites is most likely to be identified to the thermoremanent magnetization (TRM) or the partial thermoremanent magnetization (pTRM) acquired during the cooling process of these stony meteorites in the presence of a certain magnetic field, the so-called Königisberger-Thellier method, which was first introduced by KÖNIGISBERGER (1938) and later modified by THELLIER and THELLIER (1959), would be the most appropriate method to identify NRM of a stony meteorite to TRM or pTRM and evaluate its paleointensity. However, the expected linear relationship between the thermally demagnetized component of NRM, $\Delta I_n(T_o, T)$, by a thermal demagnetization from room temperature ($T_o$) to a certain higher temperature ($T$) and pTRM acquired by cooling from $T$ to $T_o$ in the presence of a magnetic field ($h$), $\Delta I_T(T_o, T, h)$, holds only for a narrow temperature range from $T_o$ to $T^*$; the systematic relationship between $\Delta I_n(T_o, T)$ and $\Delta I_T(T_o, T, h)$ does not hold any more at temperatures higher than $T^*$. In the case of Allende C3 chondrite, $T^*$ is given by $T^*=$ 150°C (BUTLER, 1972) and $T^*=130$°C (BANERJEE and HARGRAVES, 1972), and $\Delta I_n(T_o, T^*)$ is less than only 1/3 of the total NRM intensity, whence an identification of the remaining 2/3 of NRM becomes impossible.

The Königisberger-Thellier technique has been applied on a large number of Antarctic chondrites and achondrites as well as lunar rocks by one of the authors (T.N.) in a reasonably high vacuum space (about $10^{-5}$ Torr of atmospheric pressure). However, diagrams of $\Delta I_n(T_o, T)$ versus $\Delta I_T(T_o, T, h)$ for these meteoritic and lunar samples are presented by zig-zag curves at temperatures above about 150°C in most cases, thus no reliable value of the paleointensity being obtained except few exceptional meteoritic and lunar rock samples. In some such cases of lunar rock, a possible effect of the magnetic interaction between neighbouring two different magnetic phases has been suggested (*e.g.* PEARCE *et al.*, 1976), but no experimental technique to overcome such a magnetic complexity to determine the paleointensity has yet been found

It is almost certain that the Königisberger-Thellier technique, which has achieved very successful results of the paleointensity determination of terrestrial rocks, is facing significant difficulty to determine the paleointensity of lunar and meteoritic rocks, in which the principal ferromagnetic minerals are unequilibrated ferromagnetic metals consisting of Ni and small amounts of Co and P on the basis of Fe. As demonstrated by NAGATA (1979b, c, 1980b), the ferromagnetic metals in chondrites and achondrite contain, in most cases, unequilibrated martensitic plessite phase ($(\alpha+\gamma)$-phase) which is transformed to the taenite phase ($\gamma$-phase) only by heating beyond a highest transition temperature for the $(\alpha+\gamma)\rightarrow\gamma$ transition. Furthermore, it has been generally suggested that $H_2O$ contained in these extraterrestrial rocks results in various chemical reactions in heating processes, and even the oxygen gas coming from unequilibrated silicate minerals does considerably affect the composition and structure of the metallic phases in the heating process. In other words, it is practically difficult in the cases of meteoritic and lunar rocks to exactly reproduce their original cooling process from a certain high temperature above their Curie temperature in laboratories.

As an alternative method to estimate the paleointensity of NRM which is aquired by the TRM mechanism, STEPHENSON and COLLINSON (1974) have proposed to compare the AF-demagnetization curve of NRM with that of the anhysteretic remanent magnetization (ARM) acquired in the presence of a stationary magnetic field ($h$) and an alternating magnetic field decreasing from a sufficiently high value ($\tilde{H}_m$), with which the intensity of ARM in $h$ reaches the saturated value, to zero. The proposed NRM/ARM comparison method is based on the observed fact that the spectrum of microscopic coercive force of ARM against the AF-demagnetization is fairly similar to that of TRM (*e.g.* DUNLOP and WEST, 1969). Thus, a linear relation approximately holds between a part of total TRM intensity acquired in $h_T$ which is lost by the AF-demagnetization up to $\tilde{H}$, $\Delta I_T(\tilde{H}, h_T)$, and a part of ARM intensity acquired in $h_A$ which is lost by the same AF-demagnetization, $\Delta I_A(\tilde{H}, h_A)$. Namely,

$$\Delta I_T(\tilde{H}, h_T)/\Delta I_A(\tilde{H}, h_A)=f_0 h_T/h_A \qquad (1)$$

for a reasonably wide range of $\tilde{H}$. The coefficient ($f_0$) should be determined by experimentally measuring both $\Delta I_T(\tilde{H}, h_T)$ and $\Delta I_A(\tilde{H}, h_A)$. STEPHENSON and COLLISON (1974) have obtained $f_0=1.40$ for a lunar rock and NAGATA and SUGIURA (NAGATA, 1979a) have obtained $f_0=1.3$ for artificially synthesized samples of mixtures of silicates and metallic iron grains of various sizes. However, DUNLOP and WEST (1969) have shown that the $f_0$ value is sensitively dependent on ratio of the spontaneous magnetization of single-domain ferromagnetic grains, $I_s(\bar{T}_B)$, at the average blocking temperature ($\bar{T}_B$) to their spontaneous magnetization at 0°K, $I_s(0)$, their observed values of $f_0$ ranging from 1.6 to 5.0 corresponding to $I_s(0)/I_s(\bar{T}_B)$ values from 2.0 to 4.8. It seems thus that $f_0>1$ is a feasible theoretical requirement and $f_0\simeq 1.3$ will be the practical lower limit of the $f_0$-value.

Assuming that NRM is of the TRM origin and $f_0=1.3$, the paleointensity of several Antarctic achondrites has been determined with the aid of the NRM/ARM comparison method (NAGATA, 1979a, 1980a), the experimental result showing that the paleointensity ($F_p$) of the achondrites ranges from 0.01 to 0.24 Oe. Since $f_0=1.3$ is the lowest possible value of $f_0$, these evaluated values of $F_p$ of achondrites should be considered the possible upper limit of their paleointensity.

In the present study, the $f_0$-value will be experimentally determined for individual achondrites for the purpose of determining their paleointensity by the NRM/ARM comparison method.

## 2. Magnetic Properties of ALH-77302 Eucrite

ALH-77302 is a typical eucrite, NRM of which is very stable against the AF-demagnetization with respect to both the intensity and direction (NAGATA, 1980a).

The high stability of NRM is one of the main reasons why this achondrite is chosen for a test achondrite sample whose $f_0$ is specifically determined to evaluate its paleointensity. Another reason of the selection of ALH-77302 is its eucritic composition, in which the metallic component is occupied mostly by kamacite (*e.g.* NAGATA, 1980c). The other achondrites such as diogenites generally contain plessite grains as one of the main ferromagnetic metal components. The plessite phase ($\alpha+\gamma$-phase) is thermodynamically unequilibrated so that it is changed to the taenite phase ($\gamma$-phase) only by heating beyond the solvus between the $\gamma$- and ($\alpha+\gamma$)-phases (practically 520°–550°C in the case of meteorites) without any external chemical reaction.

The magnetic properties of ALH-77302 at room temperature are represented by $I_s$(saturation magnetization)=0.011 emu/gm, $I_R$(saturated IRM)=$9.3\times10^{-4}$ emu/gm, $H_C$(coercive force)=24 Oe, and $H_{RC}$(remanence coercive force)=615 Oe. In the thermomagnetic analysis, only the kamacite phase of about 750°C in $\alpha\rightarrow\gamma$ transition

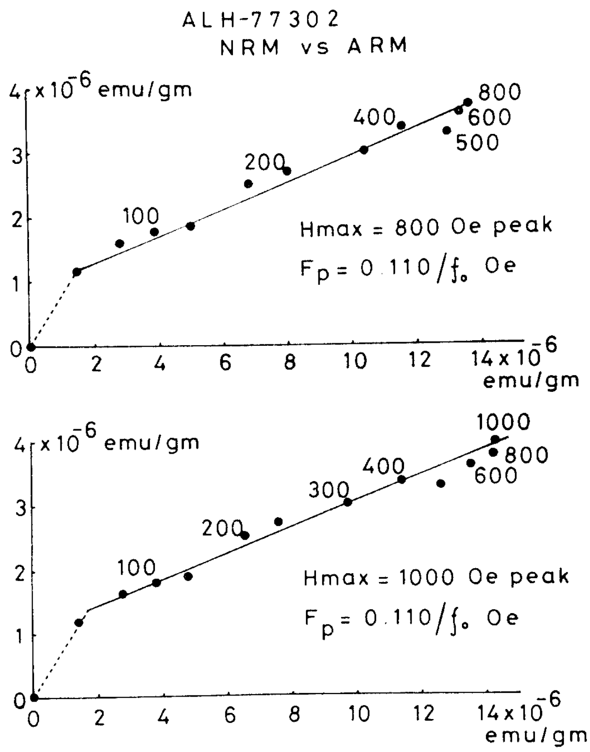


*Fig. 1. NRM-lost versus ARM-lost diagram of ALH-77302 eucrite.*
*Top: The maximum alternating magneitc field=800 Oe peak.*
*Bottom: The maximum alternating magnetic field=1000 Oe peak.*

temperature is detected in this eucrite. Then, the ferromagnetic metal constituent in this eucrite can be estimated to be only 0.055 wt% of kamacite.

The stability of NRM against the AF-demagnetization can be represented by ratio of the remaining NRM after AF-demagnetizing up to $\tilde{H}(I_n(\tilde{H}))$ to the initial NRM intensity $(I_n(0))$ for the intensity and the angular distance of the remaining NRM direction after AF-demagnetizing to $\tilde{H}$ from the original NRM before the demagnetization $(\Delta_{\tilde{H}})$ for the direction. The intensity of NRM before the AF-demagnetization of ALH-77302 is $I_n=4.14\times10^{-6}$ emu/gm, and $I_n(\tilde{H})/I_n(0)$ and $\Delta_{\tilde{H}}$ for various $\tilde{H}$ values in unit of Oe peak are given by $I_n(50)/I_n(0)=0.604$, $I_n(100)/I_n(0)=0.546$, $I_n(200)/I_n(0)=0.338$, $I_n(400)/I_n(0)=0.186$, $I_n(800)/I_n(0)=0.116$, and $\Delta_{50}=6.1°$, $\Delta_{100}=7.4°$, $\Delta_{200}=3.5°$, $\Delta_{400}=6.6°$ and $\Delta_{800}=7.4°$. It can be concluded therefore that NRM of this eucrite is very stable except for the relatively soft component of NRM which can be AF-demagnetized by $\tilde{H}=50$ Oe peak.

The AF-demagnetization curve of NRM is compared with the AF-demagnetization curves of ARM's acquired by $\tilde{H}_{\max}=800$ Oe peak and $\tilde{H}_{\max}=1000$ Oe peak and a coaxial steady magnetic field of $h=0.555$ Oe for this eucrite. Figure 1 shows the NRM-lost versus ARM-lost diagram for the AF-demagnetizations of NRM and the two cases of ARM. As shown in Fig. 1, ARM is almost saturated in $\tilde{H}=800$–1000 Oe peak, and NRM contains a soft remanent magnetization which can be AF-demagnetized by $\tilde{H}=50$ Oe peak. There is a reasonably good linear relationship between the NRM-lost and the ARM-lost after AF-demagnetizing up to 50 Oe peak. Assuming then that NRM of this eucrite is of the TRM origin, the paleointensity $(F_p)$ can be derived by eq. (1) as $F_p=0.110/f_0$ Oe.

## 3. Comparison of ARM with TRM

As already mentioned in the introduction, the main difficulty in dealing with TRM characteristics of meteorites and lunar rocks is due to considerable alterations of the ferromagnetic metal constituents in heating processes. As suggested by RIGOTTI (1978), the degree of alteration of ferromagnetic metals can be checked by comparing the acquisition characteristics of ARM and the AF-demagnetization of ARM after a heat treatment with those before the heat treatment. If there is no considerable difference between the microscopic coercivity spectra of ARM's before and after a heat treatment, characteristics of the TRM acquired by the heat treatment may well represent those of the original TRM which may correspond to NRM.

In regard to possible changes in the metallic component in lunar materials, TAYLOR (1979) has pointed out that $H_2O$, $O_2$, $N_2$ etc. along microcracks and grain boundaries of a rock sample cause serious changes in the magnetic properties of metallic component, and has proposed an effective sample preparing technique to eliminate the alteration effect in such a way as follows:

(1) Together with a metal/oxide buffer assemblage such as $FeTiO_4+FeTiO_3$

+Fe or ZnO+Zn and a getter assemblage of Ti metal, a test rock sample in a silica tube is heated at a temperature slightly higher than 100°C for a few hours, while pumping at pressure less than 10 microns $H_g$, in order to rid the sample of easily released pore water and loosely-bound contaminating gases.

(2) While still warm and attached to the vacuum line, the silica tube assembly is sealed off, where the sealed silica sample assembly consists of a part containing the sample plus the buffer and the connected other part containing the getter.

(3) The sealed tube assembly is heated at 300–400°C for an hour or two in order to make the getter absorb the tightly-bound gases, and then the silica tube containing the sample and the buffer is sealed off from the other part containing the getter.

Following the suggestion mentioned above, SUGIURA *et al.* (1979) carried out experimental approaches to reproduce the original cooling history of a lunar rock sample 70019, without putting a metal/oxide buffer. In their experimental results, the ARM characteristics can be kept practically invariant by heating up to 420°C, but they are a little changed (*e.g.* an increase of ARM by about 30%) by heating up to 590°C and considerably changed (*e.g.* an increase of ARM by about 120%) by heating up to 780°C.

It seems thus that the sample preparing technique proposed by TAYLOR for lunar materials can be applied on stony meteorites, which also contain Fe-Ni metals as the principal ferromagnetic constituent. In the present study, however, the silica capsule to seal a rock sample is not used and the metal/oxide buffer is not used as in the case of experiment by SUGIURA *et al.* Instead, the entire experimental system

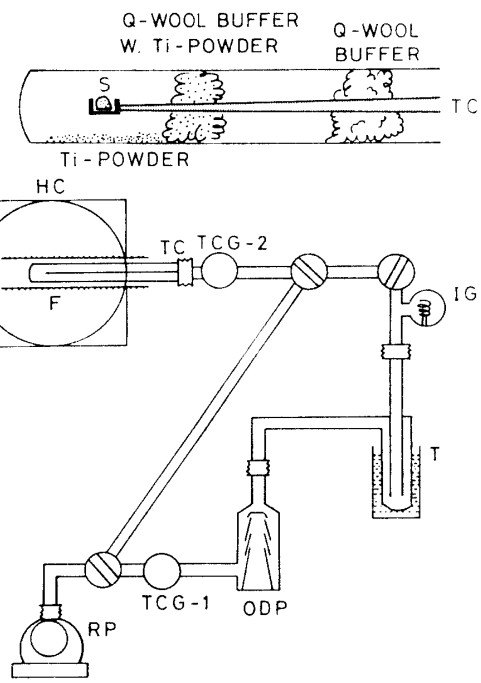


*Fig. 2. Schematic view of the thermal experiment system.*
*Top: The sample holder part*
*S: Test rock sample*
*TC: Cu-sample holder with a thermo-couple*
*Bottom: Whole system*
*HC: Helmholtz-coil system*
*F: Electric furnace*
*TC: Thermo-couple*
*TCG: Thermo-couple vacuum guage*
*IG: Ion vacuum guage*
*T: Trap*
*ODP: Oil diffusion pump*
*RP: Rotary pump*

consists of a quartz tube with a sample holder of copper along its central line and Ti-metal powder as the getter on its bottom together with quartz-wool buffers, as shown in the top of Fig. 2, and a vacuum pump system with three vacuum gages, as shown in the bottom of Fig. 2. Temperature of the sample is measured by a thermo-couple buried within the sample holder of Cu just beneath the sample. As shown in Fig. 2, a rock sample mounted on the copper sample holder within a quartz tube is horizontally set so that an electric furnace to heat the sample can be put on or taken away along the quartz tube in a Helmholtz coil system.

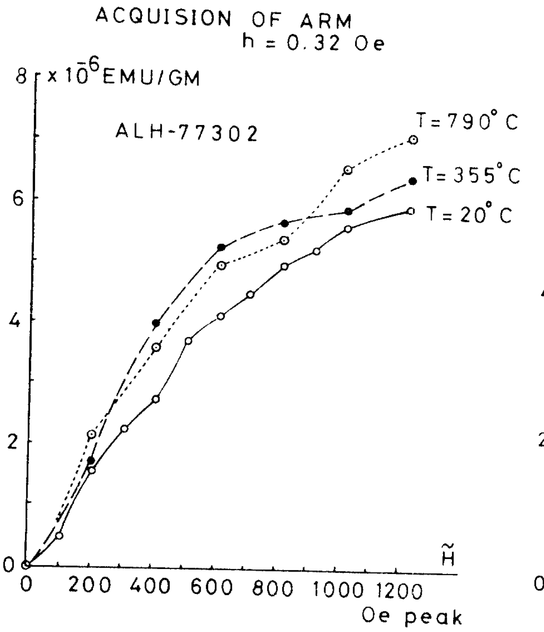


*Fig. 3. ARM acquisition curves before heating (20°C), after degasing at 355°C and after heating to 790°C ($h_A$=0.32 Oe).*

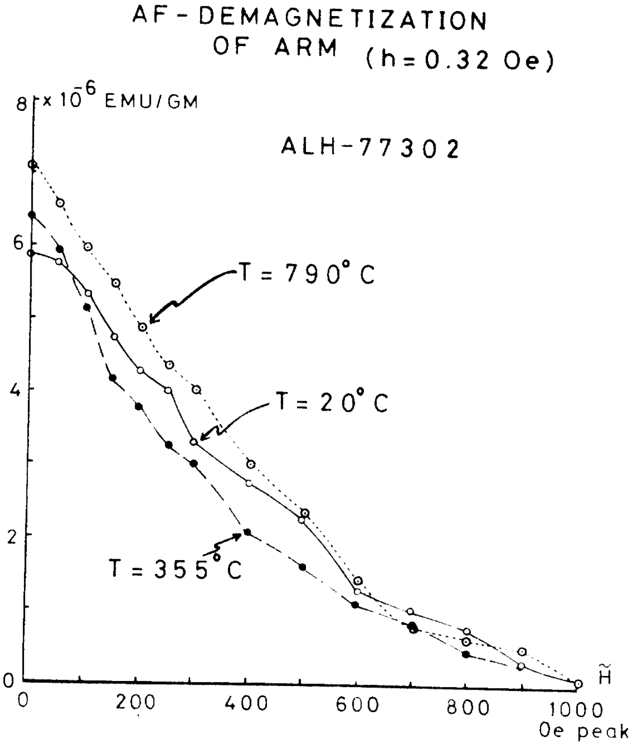


*Fig. 4. AF-demagnetization curves of ARM acquired in $h_A$=0.32 Oe and $\tilde{H}_{max}$=1220 Oe peak before heating (20°C), after degasing at 355°C and after heating to 790°C.*

In the course of a series of magnetic measurements and heating experiments on ALH-77302, firstly the stepwise acquisition of ARM in $h_A$=0.32 Oe for a $\tilde{H}$ field range from 0 to 1220 Oe peak and the AF-demagnetization of ARM acquired by $h_A$=0.32 Oe and $\tilde{H}_{max}$=1220 Oe peak are obtained at 20°C. The ARM acquisition curve and the AF-demagnetization curve of ARM at 20°C are illustrated in Fig. 3 and Fig. 4 respectively. Then, while pumping at $6\times10^{-3}$ Torr in pressure, the ALH-77302 sample is kept at 110°C for 60 min in non-magnetic space in order to rid the sample of $H_2O$ vapour and loosely-bound gases and the sample is cooled down to 20°C in non-magnetic space. After measuring the residual remanent magnetization, the sample is kept at 355°C for 90 min while continuous pumping at $6\times10^{-6}$–$1\times10^{-5}$ Torr in pressure to take out the tightly-bound gases from the sample, and then cooled down to 20°C in the same vacuum condition in the presence of a vertical magnetic field of 0.32 Oe.

After measuring the partial TRM thus acquired and the AF-demagnetization characteristics of the pTRM, the ARM acquisition and the AF-demagnetization of ARM of the heat-treated sample are examined in the same way as in the initial ARM examination. The ARM acquisition curve and the AF-demagnetization curve of the ARM are shown respectively in Figs. 3 and 4 too.

The acquisition of total TRM is carried out by cooling from 790°C in $h$=0.32 Oe in the same condition of $2\times10^{-6}$–$1\times10^{-5}$ Torr in pressure. After AF-demagnetizing the total TRM, the third examinations of ARM characteristics are carried out in the same experimental procedures as before. The ARM acquisition curve and the AF-demagnetization curve of this case also are illustrated in Fig. 3 and Fig. 4 respectively. In Fig. 3, the ARM acquisition curves of the original sample (20°C), the 355°C heat-treated sample and the 790°C heat-treated sample are in agreement with one another within 10%, and the AF-demagnetization curves of the three cases in Fig. 4 also are in agreement with one another within 10%. It may be concluded then that the present method of heating meteorite samples to acquire TRM or pTRM in high vacuum with the Ti-metal getter after taking out $H_2O$ vapour and loosely-bound gases at 110°C for a sufficiently long time appears to be successful, as far as the microscopic coercivity spectrum is concerned.

In Fig. 5, the ARM-lost values obtained by the AF-demagnetization of the initial ARM acquired in $h_A$=0.32 Oe are plotted against the total TRM-lost values obtained

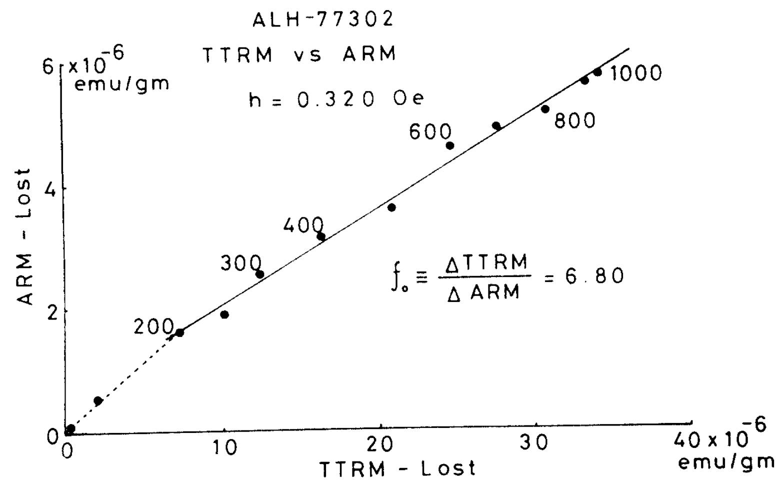


*Fig. 5. ARM-lost versus total TRM-lost diagram of ALH-77302 eucrite ($h_A=h_T$=0.32 Oe).*

by the AF-demagnetization of the total TRM thus acquired in $h_T$=0.32 Oe. The diagram indicates that the ARM-lost values are approximately proportional to the total TRM-lost values for a range of $\tilde{H}$ from 200 to 1000 Oe peak, the linear-relation coefficient being given by $f_0$=6.80.

## 4. Paleointensity of ALH-77302

In the present study, the $f_0$-value is specifically determined for ALH-77302 eucrite, assuming that NRM is entirely due to the total TRM acquired in a paleomagnetic field of $F_p$ in paleointensity. Since $F_p=0.110/f_0$ Oe in Fig. 1 and $f_0$=6.80 in Fig. 5, $F_p$ can be evaluated as $F_p$=0.016 Oe in the present case. However, the test specimen for the determination of $f_0$ is not exactly the same specimen as that for the determination of the $F_p$-$f_0$ relation, though they are closely neighbouring pieces of a same block of ALH-77302 eucrite. The NRM intensity of sample ALH-77302-B, which was used for the determination of $f_0$, increases a little with the AF-demagnetization from 0 to 300 Oe peak, and then gradually decreases with an increase of AF-demagnetization field up to 1000 Oe peak. It seems likely, therefore, that specimen ALH-77302-B possesses a less-coercive secondary NRM in addition to a high-coercive NRM. In Fig. 6, the NRM-lost values obtained by the AF-demagnetization of NRM are plotted against the TRM-lost values of total TRM acquired in $h_T$=0.32 Oe for ALH-77302-B. Assuming that the highly-coercive NRM whose NRM-lost values have a linear correlation with the total TRM-lost values represents the primary stabler NRM portion of ALH-77302 eucrite, the paleointensity can be determined by the method proposed by VAN ZIJL *et al.* (1962) as $F_p$=0.021 Oe.

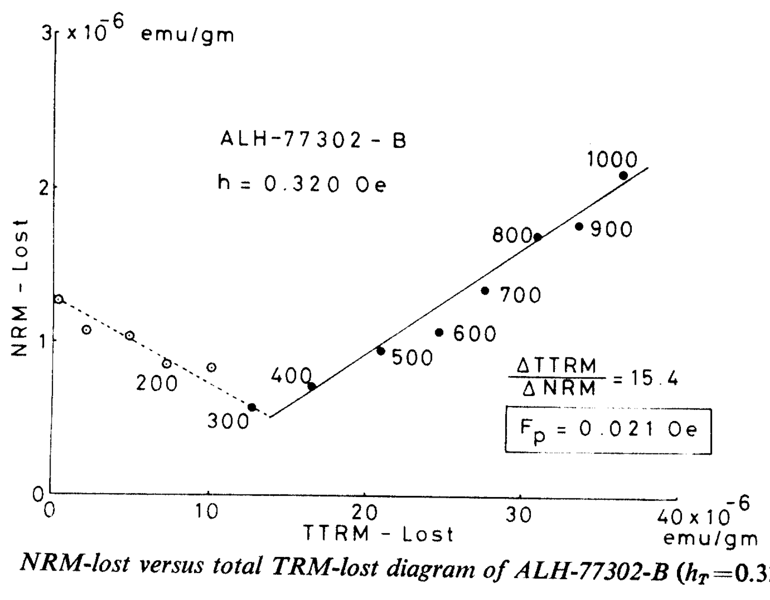


*Fig. 6. NRM-lost versus total TRM-lost diagram of ALH-77302-B ($h_T$=0.32 Oe).*

It may thus be concluded that the paleointensity of ALH-77302 eucrite is about 0.02 Oe.

## 5. Summary and Remarks

In the present work, the coefficient, $f_0$, for an achondrite is experimentally determined, and consequently the paleointensity, $F_p$, of the achondrite has become to be evaluated by comparing the AF-demagnetization characteristics of its NRM with those of its ARM.

The experimentally determined value of $f_0$ for ALH-77302 eucrite is 6.80, whereas $f_0$ determined for a lunar basalt 10053 is 1.40 (STEPHENSON and COLLINSON, 1974). On the other hand, STEPHENSON and COLLINSON (1974) and SUGIURA and NAGATA (NAGATA, 1979a) have independently examined the $f_0$-value of artificially simulated lunar or meteoritic samples, which contain a number of fine metallic iron grains uniformly dispersed within a non-magnetic matrix. Their experimental results are given by $f_0=1.28$ and $f_0=1.30$ respectively. Assuming $f_0=1.30$, then, $F_p$ of Allende C3 chondrite has been estimated by the NRM/ARM comparison method to be 0.73 Oe (NAGATA, 1979a), while $F_p$ for this C3 chondrite has been evaluated by the Königisberger-Thellier technique to be $(1.10\pm0.08)$ Oe (BUTLER, 1972; BANERJEE and HARGRAVES, 1972). Therefore, the $f_0$-value for Allende C3 chondrite may be close to unity. However, a systematic study by DUNLOP and WEST (1969) on relations between ARM and TRM of synthesized rock samples consisting of iron-oxide grains in a non-magnetic matrix has shown that the $f_0$-value is different for the ferromagnetic constituents of different chemical compositions and structures in rock samples to be examined, apparently in a positive correlation with $I_s(T=0)/I_s(\bar{T}_B)$, where $\bar{T}_B$ denotes the average blocking temperature for the TRM acquisition.

It looks likely thus that $f_0$ varies in individual lunar or meteorite rocks in general, probably within a range from 1 to 10. Since a modified Taylor technique to considerably reduce the chemical and structural alterations of metallic components at elevated temperatures during the course of heating procedure is ascertained in the present work, it will be suggested that $f_0$ should be determined for individual meteoritic samples with the aid of the present technique or a similar one in the future studies of the paleointensity of stony meteorites.

The paleointensity of achondrites reported in the previous paper (NAGATA, 1980c) has been derived from the NRM/ARM comparison experiments on an assumption of $f_0=1.3$. As stated in the paper, the estimated values of $F_p$ (0.010–0.089 Oe) of those achondrites present their possible upper limits. Therefore, the conclusion given in the previous paper that "the upper limit of paleointensity of most achondrites is 0.1 Oe or less" will still be able to stand.

In concluding, the writers' thanks are due to M. D. FULLER for his cooperation and advice for the present experimental works.

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